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OCEANIC ROLE IN TERRESTRIAL CLIMATE

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ABSTRACT

A review is given of the role of the oceans in terrestrial climatogenesis. The review leads to development of a criterion for recognizing areas of potential climatogenesis through atmosphere-oceanic interaction and to speculations as to the causes of the El Ninõ-Southern Oscillation events, the Medieval Little Optimum, Little Ice Age and 20th Century Arctic Warming of the North Atlantic and the glacial-interglacial cycles.

1. INTRODUCTION

Somewhat more than half of the received solar energy is absorbed at the planetary surface. The oceans, because of their greater area (~71% of the global surface), lower albedo (0.05–0.10) and preponderance in low latitudes (~76% of the area from 20N to 20S), account for over three-quarters of the surface absorption and thus approximately 55% of the solar energy absorbed by the Earth-atmosphere system. Since only about 10% of this is reradiated directly to space, approximately 50% of the solar energy absorbed by the Earth-atmosphere system is delivered to the atmosphere in the forms of sensible and latent heat and thermal radiation from the underlying ocean surface. The times, locations, forms and rates of delivery of this energy to the atmosphere are determined principally by the ocean surface temperature, the surface wind speed and the temperature and moisture gradients between the boundary layer of the atmosphere and the underlying ocean surface. This assures that the oceans play a dominant role in terrestrial climate.

In addition to the thermal rôle, watér evaporated principally from the oceans returns as rainfall, averaging about 1 m per year globally. The river run-off to the oceans tells us how much of the precipitation was recently evaporated from the oceans – ultimately, it all was. Note that this role also makes possible the cryosphere, consisting of permafrost, snow cover, land glaciers and sea ice. While each of these leads to significant climatogenic atmospheric-oceanic interactions, with feedback loops on many time scales, there is not space to consider them here.

2. STATISTICALLY OBVIOUS OCEANIC CLIMATOGENESIS

A part of the climatogenic role of the oceans can be seen in Fig. 1—showing the annual average incoming solar minus outgoing IR or net radiation. This clearly shows that the Earth receives more radiation than it emits in low latitudes and radiates away more radiation than it receives in high latitudes. This is only possible if the low latitude excess is somehow transported to higher latitudes within the system. The transport is performed, of course, by both the atmosphere and the oceans—roughly fifty-fifty by current estimates. Note, however, that the zero (thickened) line is displaced poleward over the oceans and equatorward over the continents. This shows immediately that the presence of the oceans causes our planet to be warmer than it otherwise would be.

The tremendous impact of the oceans as a thermoregulator of terrestrial climate is more apparent in Fig. 2--showing a map of the seasonal range in surface *air* temperature. This map does not show continental outlines but the locations of the individual continents are readily identifiable. Only over and in proximity to the continents and Arctic sea ice does the seasonal range in air temperature exceed 12°C—over Asia it reaches 56°C, North America 44, Antarctica 36, North Africa 24, and elsewhere, no more than 16.

When I first examined a map of this quantity, I was surprised by the bands of relative minimum range over the oceans circa 50–55° latitude—present in both hemispheres but far more obvious in the SH (Southern Hemisphere). However, their explanation is clear from Fig. 3 which shows the seasonal range in surface *water* temperature or SST (sea surface temperature). SST has a seasonal range of 1–2°C at the equator, which increases poleward to bands of maximum range in mid-latitudes. The maximum seasonal range is both greater and somewhat farther poleward in the NH (7–15°C at 40–50°N) than in the

SH (4–8°C at 30–45°S). Poleward of these bands the seasonal range in SST again decreases toward the data void areas at latitudes greater than 50–55°. A little reflection reveals that the seasonal range in surface *water* temperature must vanish at the summer ice boundary, since at each point on, and poleward, of this boundary the surface water is constrained to remain at the freezing point of the sea ice with which it is in contact throughout the year. Boundary layer air blowing off the continents is soon brought close to equilibrium with the SST—so is constrained to follow the seasonal range in SST far from land or sea ice. This accounts for the bands of minimum seasonal range of surface *air* temperature shown on Fig. 2 near 50–55°.

Figure 4 shows another example of oceanic control of climate. This is a map of the vector standard deviation of the 500-mb wind in winter. Kinetic energy generation and dissipation are roughly proportional to the cube of this quantity (Ellsaesser, 1969). The centers of maximum over the oceans are indicative of centers of maximum transformation of potential to kinetic energy by cyclogenesis. Cyclogenesis is favored in these areas by the low frictional drag of the ocean surface, by the strong thermal gradients developing along the downwind edge of nontropical continents and by the heat and moisture available from the warm western boundary currents of the oceans.

3. OPERATING MODES OF THERMOREGULATORY CONTROL

The low diurnal and seasonal range of SST gives evidence of the strong thermoregulatory control exerted by the oceans on terrestrial climate. This thermoregulatory control is quite complex and includes the capability, through atmosphere-ocean interaction, of changing the timing, locations and modes of energy release to the atmosphere and transfers from one part of the atmosphere to another.

3.1 *Thermal Reservoir Modes*

A large degree of thermoregulatory control is provided simply by the large heat capacity of the mixed layer of the ocean in thermal contact with the atmosphere. This layer has an average thickness of about 70 m. Seventy meters of water has almost 30 times the heat capacity of the atmospheric column above it. At the average rate of absorption of solar energy at the earth's surface it would take almost 20 days to receive enough energy to warm such a layer 1°C. The oceans provide even larger thermal reservoirs against cooling in high latitudes and warming in low latitudes. Near the poles, ocean surface cooling is reduced first by the release of latent heat through the freezing of sea water. Then, once an ice cover forms, evaporation is greatly reduced and heat lost is reduced by the low thermal conductivity of the ice which allows the upper surface of the ice to cool significantly below the temperature of the water below and thus to reduce the rate of loss of radiative energy to the overlying atmosphere and to space. As the SST increases in lower latitudes a larger and larger fraction of the absorbed radiation (both solar and back IR from the atmosphere) is used to evaporate water rather than to raise its temperature. Under normally observed conditions, at ocean surface temperatures above ~30°C, evaporation occurs at such a rate that it begins to draw sensible heat out of the atmosphere (Ellsaesser, 1984).

3.2 *The Conveyer Belt Mode*

Radiant energy, both solar and back IR from the atmosphere, absorbed by the mixed layer of the ocean is transported elsewhere by the atmospherically driven ocean currents—generally poleward and toward the eastern boundary of the ocean—before the energy is released back to the atmosphere. This effect accounts, for example, for the mild climate of the northeastern North Atlantic. This mode is presumably controlled by the strength, scale and constancy of the general circulation of the atmosphere. Similar but weaker cold currents occur along the eastern boundaries of the oceans.

3.3 *The ITCZ-Hadley Cell Mode*

Boundary layer air, warmed and moistened by contact with the tropical oceans, is swept by the trade winds into narrow bands of deep convection paralleling the equator—the so-called ITCZ (Intertropical Convergence Zones). Here both sensible and latent energy are quickly removed from the surface, carried up through the moist layer and spread laterally above most of the greenhouse effect of atmospheric water vapor, where the energy can more readily be reradiated to space. The two-dimensional Hadley circulation, thermally driven by the ITCZ convection, carries potential energy poleward where it is converted to sensible heat by later subsidence. This serves several thermoregulatory functions. It removes energy from the surface in the tropics at a much lower temperature than would be required to eject it radiatively, it transports heat from low to high latitudes and over the tropics it significantly reduces the effectiveness of our principal greenhouse gas, water vapor. It reduces the water vapor greenhouse effect in two ways—it first circumvents it by physically transporting sensible and latent energy up through and spreading it horizontally above the moist layer; and secondly it reduces the greenhouse effect directly by thinning the moist layer by the large scale subsidence which completes the downdraft legs of the

Hadley circulation. This thermoregulatory mode is presumably controlled, not by the strength per se, but by the rate of convergence of the two trade wind systems. Due to the rapid increase of water vapor tension with temperature, it must also depend on tropical ocean surface temperature. However, this effect may be inverse- –since the greater the energy density of the rising currents, the smaller the volume required to remove energy from the surface at the constant rate at which it is received from the sun.

3.4 *The Tropical Cyclone Mode*

In this mode the warm moist boundary air over the tropical oceans is swept into converging spiral bands of convection which can form only over the low friction surface of the oceans, and displaced more than 5° latitude from the equator. The rapid transfer of latent and sensible heat through, and its spreading above, the moist layer occurs much the same as in ITCZ convection. However, the surrounding subsidence lacks zonal symmetry and presumably does not contribute to the Hadley circulation. On the other hand, since most tropical cyclones recurve into the westerlies, each carries a large pulse of both sensible and latent heat into higher latitudes. Thus, tropical cyclones accomplish all the same thermoregulatory actions as the Hadley circulation. In addition, they stir the mixed layer of the ocean, mixing heat down into the thermocline and bringing cooler water to the surface. This mode is presumably controlled by the degree to which the ocean surface temperature exceeds the 27°C threshold for tropical cyclone formation, and the depth, and the absence of vertical shear in the convectively unstable trade wind currents.

3.5 *The Thermohaline Circulation Modes*

While the thermohaline circulation is planetary in scale, it is useful to think of it as having two distinct circulation cells; one which has a more-or-less globally uniform upwelling, and a separate one in which the upwelling is restricted to the areas of recognized surface upwelling. In both cases the downwelling (or bottom water formation) occurs where there is rapid cooling of saline surface water and/or where dense water is produced when salt is forced out of newly formed sea ice in the North Atlantic and around Antarctica.

In the *Uniform Thermohaline Mode* the entire oceanic volume is filled by dense cold water sinking to the bottom near the winter poles. As newer bottom water is added, it forces the older water uniformly upward and, where this meets heat diffusing down from the sun-warmed surface mixed layer, is created the strong vertical temperature gradient identified as the thermocline. The loop of course is closed by surface flows of warm water toward the winter poles. This mode acts to transfer cold water from high to low and warm water from low to high latitudes and thus to moderate extremes of cold in high latitudes and heat in low latitudes. It is controlled by the rate of formation of bottom water which in turn would appear to be related to the volume of sea ice formed each year.

The *Local Thermohaline Mode* accomplishes the same role except that the cold water is identifiable where it returns to the surface in lower latitudes in favored locations recognized as areas of surface upwelling. As an example of this mode we will consider only the upwelling along the equator in the eastern Pacific and down along the coast of South America— the El Niño area. The El Niño phenomenon is part of the overall El Niño-Southern Oscillation event. The east-west see-saw of surface pressure in the equatorial and southern Pacific has been recognized since 1897 (Rasmussen and Carpenter, 1982), but it was Bjerknes (1966) who linked it with the interannual SST fluctuations of the El Niño. The entire event is a multi-year oscillation lasting 3 or more years. During each El Niño, the austral summer maximum in SST at Puerto Chicama, Peru rises above its normal maximum of 19°C by 1 to 4.5°C for 2 or 3 consecutive years (Wyrkti, 1975). In the usual case of 2 years, one of the warmings predominates and the second occurs 2 to 3 months earlier than usual. However, some authors have reported periods of 1 and 2 years between El Niños, indicating nearly equivalent warmings in consecutive or alternate years of the cycle.

SST warmings extending northward from Peru to the equator and then westward to the dateline are now considered a more significant aspect of the El Niño. In this area, SST is normally cooled by upwelling by up to 6°C below that normally found at corresponding latitudes. Among other effects, this provides a thermal reservoir capable of accepting and storing a large amount of solar radiation—i.e., received solar radiation is converted into thermal energy remaining in the mixed layer rather than being converted into latent heat that is released into the overlying atmosphere, as is done where the SST is at its normal equatorial value (Csanady, 1984). This heat can then be carried in the surface ocean currents to be released months to years later and thousands of kilometers from the point at which received. Occurrence of the El Niño interrupts this mode, and causes the received solar energy to be returned with little delay to the overlying atmosphere, primarily as latent heat of water vapor (Ellsaesser, 1984; Csanady, 1984). This appears to enter and strengthen the Hadley circulation. That is, a pseudo-steady state is interrupted by a circulation mode which for 12–18 months accelerates the release of sensible heat over that portion of the atmosphere affected by the Hadley circulation originating from the El Niño area. However, this accelerated release of energy is largely injected above the greenhouse effect of water vapor where it can easily be radiated to space, and it is followed by periods of decelerated release, as solar energy is again stored in cold surface water as the normal upwelling is renewed.

Remember that the solar energy that would have normally been stored in the upwelling cold water during the time of the El Niño was promptly lost from the ocean and is no longer available for return to the atmosphere during this part of the cycle. This seems to fit the studies of Angell and Korshover (1978a,b,1983) which detect apparent warm pulses followed by cool pulses emanating from the El Niño area and propagating poleward and upward throughout most of the troposphere and stratosphere. Beyond the area of the SST anomaly itself, the cool pulses are the more apparent. (Note: It appears highly possible that such El Niño cooling pulses have in the past been mistakenly attributed to the effect of stratospheric clouds created by strong volcanic eruptions (Ellsaesser, 1983). The El Niño years since 1868 cited by Quinn *et al.* (1978) show a high degree of overlap with volcanic eruptions listed by Simkin *et al.* (1981).)

This local thermohaline mode is presumably controlled by the strength of the surface winds driving the oceanic upwelling in the regions in question. For the El Niño, the strength of the southeast trades in the South Pacific would appear to be dominant. The CLIMAP (1976, 1981) studies tend to confirm this since they found SST cooling in the area significantly more extensive 18,000 BP (before present) when the zonal winds in the atmosphere were presumably stronger than now.

4. POTENTIAL OCEANIC CLIMATOGENESIS

A global map of average January SST isotherms is shown in Fig. 5. Note that in most areas these isotherms are nearly parallel to latitude circles. This is, the pattern to be expected if SST were determined by processes other than advection under a pseudo-zonally uniform atmosphere. When the SST isolines depart from this pattern we can be sure that oceanic advection of some type is involved, and since oceanic currents are driven primarily by atmospheric winds we can be equally certain that they are determined by some type of atmosphere-oceanic interaction. Since even a slight change in one partner of such interactions can lead to large changes before new equilibria are established, each area in which SST isotherms depart significantly from parallel east-west lines should be regarded as an area of potential climatogenesis. The two regions of Fig. 5 showing the most marked departures from zonal SST isolines are the North Atlantic and the eastern South Pacific. Both are areas already diagnosed as having significant climatogenetic atmosphere-oceanic interaction. The El Niño-Southern Oscillation of the tropical Pacific has already been mentioned but it warrants further speculation.

It is well known that the atmosphere gains angular momentum from the nongaseous Earth wherever there are easterly winds at the surface and returns it wherever there are surface westerlies. Also, the globally averaged angular momentum of the atmosphere undergoes an annual cycle with a maximum of westerly angular momentum in the boreal winter and a minimum in the austral winter (Rosen and Salstein, 1983), due to domination by the seasonal cycle in winds of the NH. Due to the many other physical constraints on the atmosphere, it would not be surprising if the atmosphere were unable to maintain at all times the areas and strengths of surface easterlies and westerlies required to keep angular momentum flowing at just the right rate to follow its seasonal cycle in perfect "equilibrium."

It seems plausible, that as the atmosphere strengthens and expands the area of surface easterlies to build up to the January maximum of angular momentum, it could overshoot and build up so much westerly angular momentum that it then has difficulty maintaining surface easterlies. As the westerly angular momentum is moved into higher latitudes and altitudes and the easterlies weaken, they drop to the point that the trade-wind-driven equatorial upwelling in the eastern Pacific cannot be maintained. Without the westward surface drag and resultant Coriolis-imposed divergence at the surface, both solar heating and hydrostatic readjustment within the ocean will cause the surface water to warm. Because it is the eastward component of the trades that is reduced (the equatorward component is actually increased (Quiroz, 1983)), the warm SST leads to enhanced penetrative convection—carrying air with little or no easterly component—actual westerly components in the western Pacific (Pazan and Meyers, 1982)—into the upper atmosphere. This represents a substantial addition of westerly momentum to the upper atmosphere compared to the air usually ascending in tropical convection. Here is a possible positive feedback mechanism, by which an atmosphere, with already excessive westerly momentum, develops an additional mechanism for generating more. There of course remains the question as to whether the enhanced convection and Hadley circulation generate as much westerly momentum as would have the normal easterlies, whose weakening or absence brought the former into being. Suggestive confirmation is indicated by the tendency of the El Niño, once set in motion, to continue through at least 2 austral summers, and for the peak SST anomaly to be accompanied by a peak positive anomaly in the westerly angular momentum of the atmosphere (Stefanick, 1982; Rosen *et al.*, 1984).

Thus, the El Niño-Southern Oscillation appears to be an amplification of the normal annual cycle of angular momentum with a positive feedback mechanism providing an additional source of westerly angular momentum peaking in the austral summer. It appears to be initiated, or at least preceded by a year or more of stronger than normal easterly trades giving higher sea level in the western Pacific and cooler SSTs due to stronger upwelling (Wyrtki, 1975). This should show up as a negative anomaly in westerly angular momentum of the atmosphere and in turn lead to an above normal rate of gain of westerly angular momentum from the Earth. What terminates the self-amplifying cycle isn't clear but the available data suggest two possibilities. First, in most El Niños for which we have data, the second amplified austral summer warming of SST off Peru occurs 1–3 months earlier than normal. This could allow the normal processes which build the austral summer southeast trades of the South Pacific to redevelop normally to bring the SSTs back to normal through upwelling. Secondly, the peak SST anomalies of the El Niño, and particularly that of the unusual El Niño of 1982–83 were accompanied by strong positive anomalies in westerly angular momentum of the atmosphere (Stefanick, 1982; Rosen *et al.*, 1984). Once this works its way back to the surface, frictional drag at the earth's surface will also be above normal and could bring the atmosphere back to its normal "equilibrium" position which again allows the easterly trades to develop normally and cool the SST by renewed upwelling.

Atmosphere-ocean interactions in the North Atlantic appear both more probable and of more far reaching consequences. We have heard of the North Atlantic see-saw since the time of Walker but this does not appear to be a very large perturbation. Of far greater consequences were the oscillations known as the Medieval Little Optimum and Little Ice Age, leading to the colonization of Greenland and its later abandonment, and the dramatic warming of the Arctic and northern North Atlantic from 1918 to 1938 known earlier as The Arctic Warming. These remain unexplained but are reasonable expectations for a slight shift in the North Atlantic polar front, the region of isotherm packing in the eastern North Atlantic in Fig. 5.

Better documented are the shifts in this front since the peak of the last glacial 18,000 BP. Figure 6 shows the most recent positions as determined by Ruddiman and McIntyre (1981) from cores of sediment taken from the bottom of the North Atlantic. According to these studies the North Atlantic polar front swung like a gate hinged southeast of Newfoundland. Also, the retreat from the southernmost position at the peak of the last glacial was not monotonic—from 13,000 to 11,000 BP the eastern end of the front readvanced thousands of kilometers to the south and then retreated again between 10,000 and 9,000 BP. This period of glacial readvance is well documented as the Younger Dryas cold period of northwestern Europe and Scandanavia (Ruddiman and McIntyre, 1981).

If such swings of the North Atlantic polar front are possible in as short a period as 1,000 to 2,000 years, then smaller swings may account for the dramatic climate changes in this area now called the Medieval Little Optimum, The Little Ice Age and The Arctic Warming. It is quite reasonable to expect that even slight shifts in the position of the North Atlantic polar front of the type shown in Fig. 6 would lead to significant changes in mean annual temperatures over large areas of the North Atlantic, Eurasia and even North America. From this and the paucity of evidence of any significant change in the volume of Antarctic glaciers (Burckle *et al.*, 1982), it also appears plausible that the entire glacial-interglacial cycle could have occurred simply as a result of changes in the position of the North Atlantic polar front. What is not so obvious, of course, is whether changes in the polar front position were the cause, or even preceded, the climatic changes. Ruddiman and McIntyre (1981) concluded that the bulk of the melt water from the Laurentide deglaciation occurred prior to 12,000 BP and thus before the final retreat of the polar front across the North Atlantic. A fundamental approach would appear to be to try to understand what physical processes cause the packed isotherms off Nova Scotia in Fig. 5 to bend sharply left into the Labrador Sea.

The SST isotherms of Fig. 5 also show a large area of departure from an east-west orientation in the South Atlantic suggesting the potential for interactive climatogenesis there as well. The fact that no significant climatic fluctuations have been reported for this area may be due simply to insufficient observational data or it might be due to the fact that the relatively narrow width of the South Atlantic does not allow sufficient variation in position or intensity of the anticyclone trapped there.

SUMMARY

The oceans constitute a robust and multimodal thermoregulator of planetary surface temperature. In addition to the obvious thermal flywheel of the mixed layer averaging ~70 m in depth, (i) the latent heat of evaporation limits tropical temperature, (ii) both the latent heat of freezing and the insulation effect of sea ice limit planetary energy loss in high latitudes, and (iii) both wind driven and thermohaline circulations transport warm water from low to high and cold water from high to low latitudes. Both the low frictional drag of the ocean surface and the inexhaustible supply of water for evaporation support the ITCZ-Hadley circulation and tropical cyclones which efficiently circumvent over the tropics our principal greenhouse gas, water vapor, by removing energy from the surface in low latitudes, convecting it through and spreading it horizontally above the moist layer for easier reradiation to space. In addition, these circulations transport energy to higher latitudes and directly reduce the greenhouse effect of water vapor by the drying effect of the broadscale subsidence which completes the downward legs of these convectively driven circulation cells.

It is pointed out that departures of SST isotherms from parallel east-west lines show evidence of atmospheric-oceanic interactive transport and thus indicate areas of potential climatogenesis, since a slight perturbation can lead to large changes before new equilibria are established. The North Atlantic and the eastern tropical Pacific are the most outstanding areas in which this is observed. Important climatic fluctuations have been observed in both areas; El Niño in the Pacific and the Medieval Little Optimum, Little Ice Age and The Arctic Warming of this century in the North Atlantic. Speculations are given as to the causes of these climatic fluctuations and the glacial-interglacial cycle.

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REFERENCES

Angell, J. K. and J. Korshover, "Global Temperature Variations in the Troposphere and Stratosphere, 1958–1982," *Mon. Wea. Rev., 111*, 901–921, 1983.

Angell, J. K. and J. Korshover, "Global Temperature Variation, Surface-100 mb: An Update into 1977," *Mon. Wea. Rev., 106*, 755–770, 1978a

Angell, J. K. and J. Korshover, "Estimate of Global Temperature Variations in the 100-30 mb Layer Between 1958 and 1977," *Mon. Wea. Rev., 106*, 1422–1432, 1978b.

Bjerknes, J., "A Possible Response of the Atmospheric Hadley Circulation to Equatorial Anomalies of Ocean Temperature," *Tellus, 18*, 820–829, 1966.

Bjerknes, J., "El Niño Study Based on Analyses of Ocean Surface Temperatures, 1935–1957," *Bull. Inter-Amer. Trop. Tuna Com., 5*, 217–303, 1961.

Burckle, L. H., D. Robinson and D. Cooke, "Reappraisal of Sea-ice Distribution in Atlantic and Pacific Sectors of the Southern Ocean at 18,000 yr BP," *Nature, 299*, 435–437, 1982.

CLIMAP Project Members, "The Surface of the Ice-age Earth," *Science, 191*, 1131–1137, 1976.

CLIMAP Project Members, "Seasonal Reconstructions of the Earth's Surface at the Last Glacial Maximum," *Geol. Soc. Am.* Map and Chart Series No. 36, 1981.

Crutcher, H. L., "Upper Wind Statistics Charts of the Northern Hemisphere," *NAVAER 50-1C-535*, Vol. 1, Office of the Chief of Naval Operations, Ashville, NC, 1959.

Csanady, G. T., "Warm Water Mass Formation," *J. Phys. Oceanogr., 14*, 264–275, 1984.

Ellsaesser, H. W., "The Climatic Effect of CO_2: A Different View," *Atmos. Environ., 18*(2), 431–434, 1984.

Ellsaesser, H. W., "Isolating the Climatogenic Effects of Volcanoes," Lawrence Livermore National Laboratory Report UCRL-89161, Livermore, CA 94550, 1983.

Ellsaesser, H. W., "A Climatology of Epsilon (Atmospheric Dissipation)," *Mon. Wea. Rev., 97*, 415–423, 1969.

Levitus, S., *Climatological Atlas of the World Ocean*, NOAA Professional Paper 13, U. S. Government Printing Office, Washington, DC 20402, 1982.

Luther, D. S., D. E. Harrison and R. A. Knox, "Zonal Winds in the Central Equatorial Pacific and El Niño," *Science, 222*, 327–330, 1983.

Monin, A. S., "The Role of the Oceans in Climatic Models," pp. 201–205 in *The Physical Basis of Climate and Climate Modeling*, GARP Publ. No. 16, WMO-ICSU, 1975.

Panfilova, S. G., "Seasonal Surface Water Temperature Variations in the World Ocean," *USSR Acad. Sci. Oceanology, 12*(3), 333–344, 1972.

Pazan, S. E. and G. Meyers, "Interannual Fluctuations of the Tropical Pacific Wind Field and the Southern Oscillation," *Mon. Wea. Rev., 110*, 587–600, 1982.

Quinn, W. H., D. O. Zopf, K. S. Short and R. T. W. Kuo Yang, "Historical Trends and Statistics of the Southern Oscillation, El Niño, and Indonesian Droughts," *Fishery Bulletin, 76*(3), 663–678, 1978.

Quiroz, R. S., "The Climate of the 'El Niño' Winter of 1982–83—A Season of Extraordinary Climatic Anomalies," *Mon. Wea. Rev., 111*, 1685–1706, 1983.

Rasmusson, E. R. and T. H. Carpenter, "Variations in Tropical Sea Surface Temepratue and Surface Wind Fields Associated with the Southern Oscillation/El Niño, *Mon. Wea. Rev., 110*, 354–384, 1982.

Rosen, R. D. and D. A. Salstein, "Variations in Atmospheric Angular Momentum on Global and Regional Scales and the Length of Day," *J. Geophys. Res., 88*, 5451–5470, 1983.

Rosen, R. D., D. A. Salstein, T. M. Eubanks, J. O. Dickey and J. A. Steppe, "An El Niño Signal in Atmospheric Angular Momentum and Earth Rotation," *Science, 225*, 411–414, 1984.

Ruddiman, W. F. and A. McIntyre, "The North Atlantic Ocean During the Last Deglaciation," *Palaeogr., Palaeocli., Palaeoeco., 35*, 145–214, 1981.

Simkin, T., L. Seibert, L. McClelland, D. Bridge, C. Newhall and J. H. Latter, *Volcanoes of the World*, Hutchinson Ross, Stroudburg, PA, 1981.

Stefanick, M., "Interannual Atmospheric Angular Momentum Variability 1963- 1973 and the Southern Oscillation," *J. Geophys. Res., 87*, 428–432, 1982.

Stephens, G. L., G. G. Campbell and T. H. Vonder Haar, "Earth Radiation Budgets," *J. Geophys. Res., 86*, 9739–9760, 1981.

Walker, G. T., "World Weather IX," *Mem. Indian Meteor. Dept., 24*, 275–332, 1924.

Wyrtki, K., "El Niño, The Dynamic Response of the Equatorial Pacific Ocean to Atmospheric Forcing," *J. Phys. Oceanogr., 5*, 572–584, 1975.

Legends for Figures

FIGURE 1. Map of mean annual net radiation at the top of the atmosphere [W/m^2] (from Stephens *et al.*, 1981).

FIGURE 2. Map of seasonal range in surface air temperature [°C] (adapted from Monin, 1975).

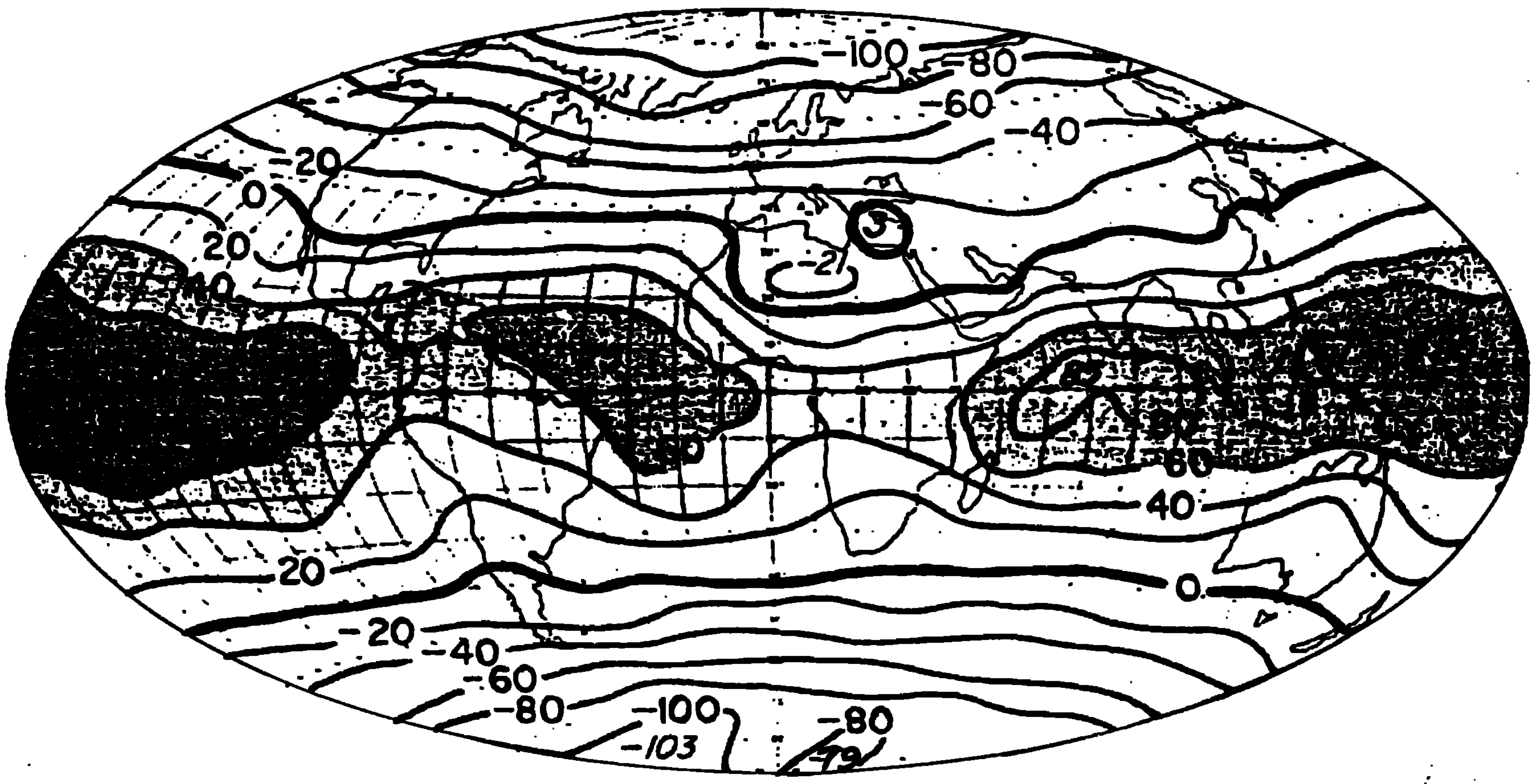
FIGURE 3. Map of seasonal range in SST (sea surface temperatue)[°C] (adapted from Panfilova, 1972).

FIGURE 4. Vector standard deviation of the 500-mb wind in winter [knots] (adapted from Crutcher, 1959).

FIGURE 5. Global map of average SST isotherms for the month of January [°C] (from Levitus, 1982.

FIGURE 6. Locations of the polar front of the North Atlantic at various times during the most recent deglaciation as determined from ocean sediment cores (adapted from Ruddiman and McIntyre, 1981).

ANNUAL NET [W/m²]



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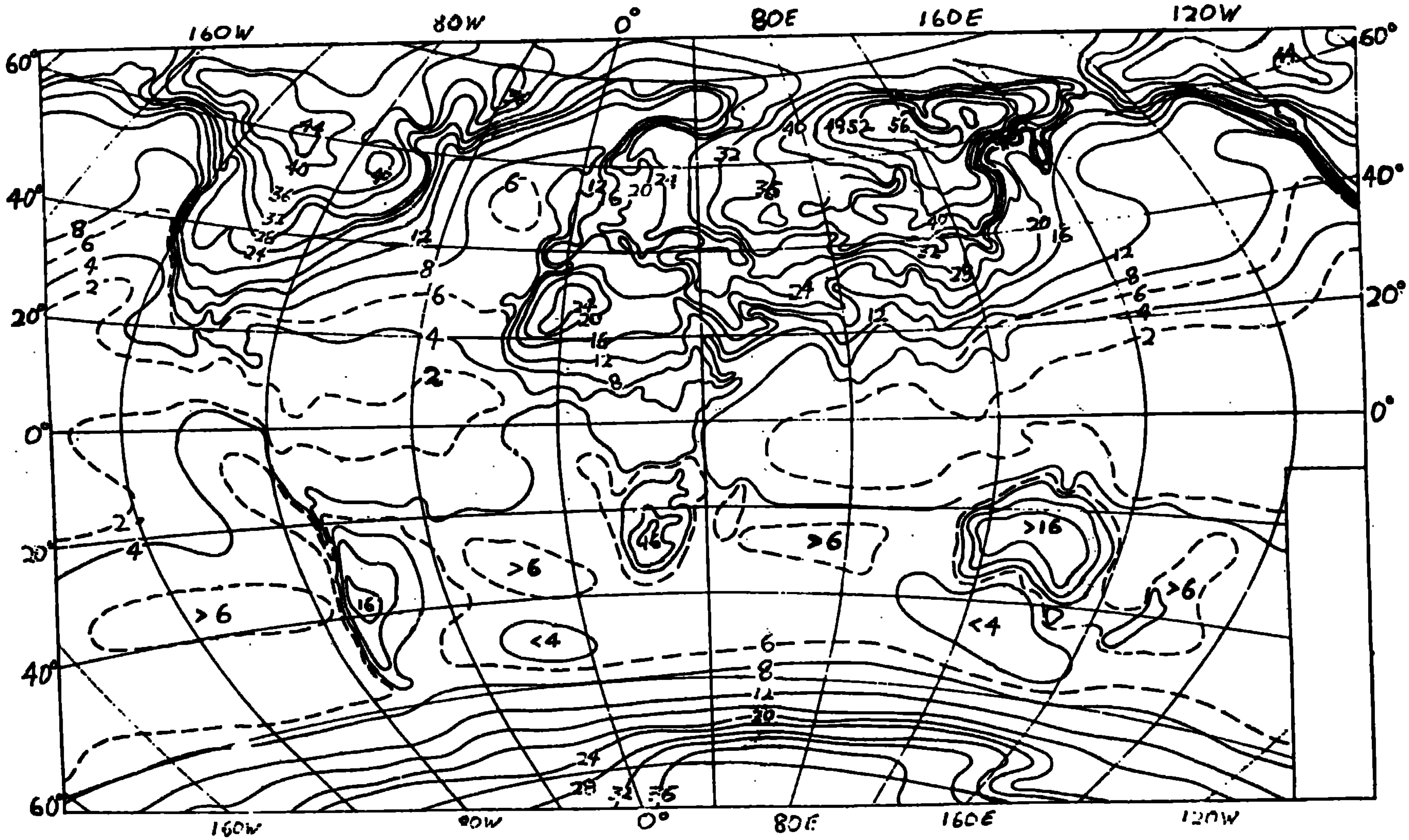

FIGURE 2. Map of seasonal range in surface air temperature [°C] (adapted from Monin, 1975).

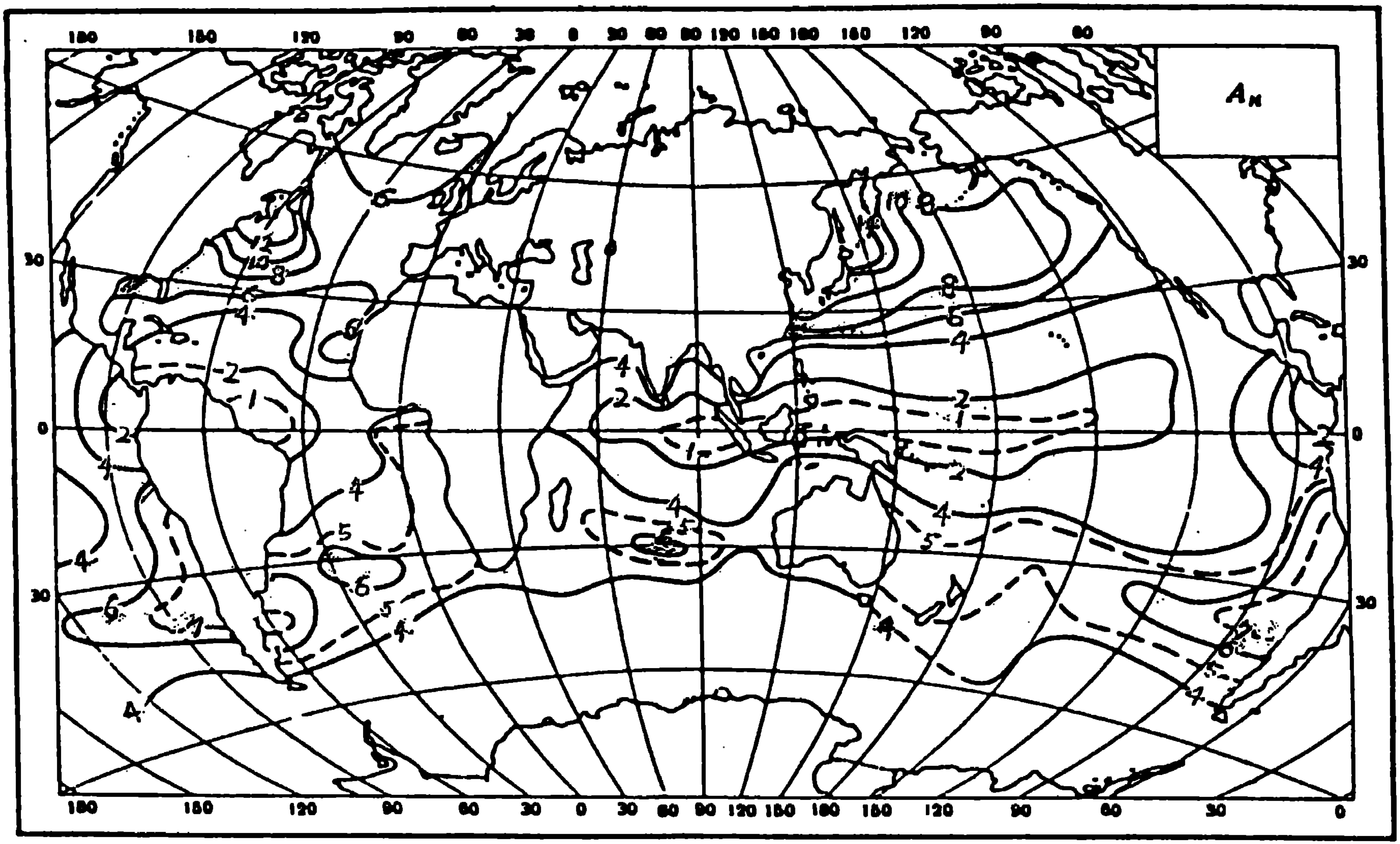

FIGURE 3. Map of seasonal range in SST (sea surface temperatue)[°C] (adapted from Panfilova, 1972).

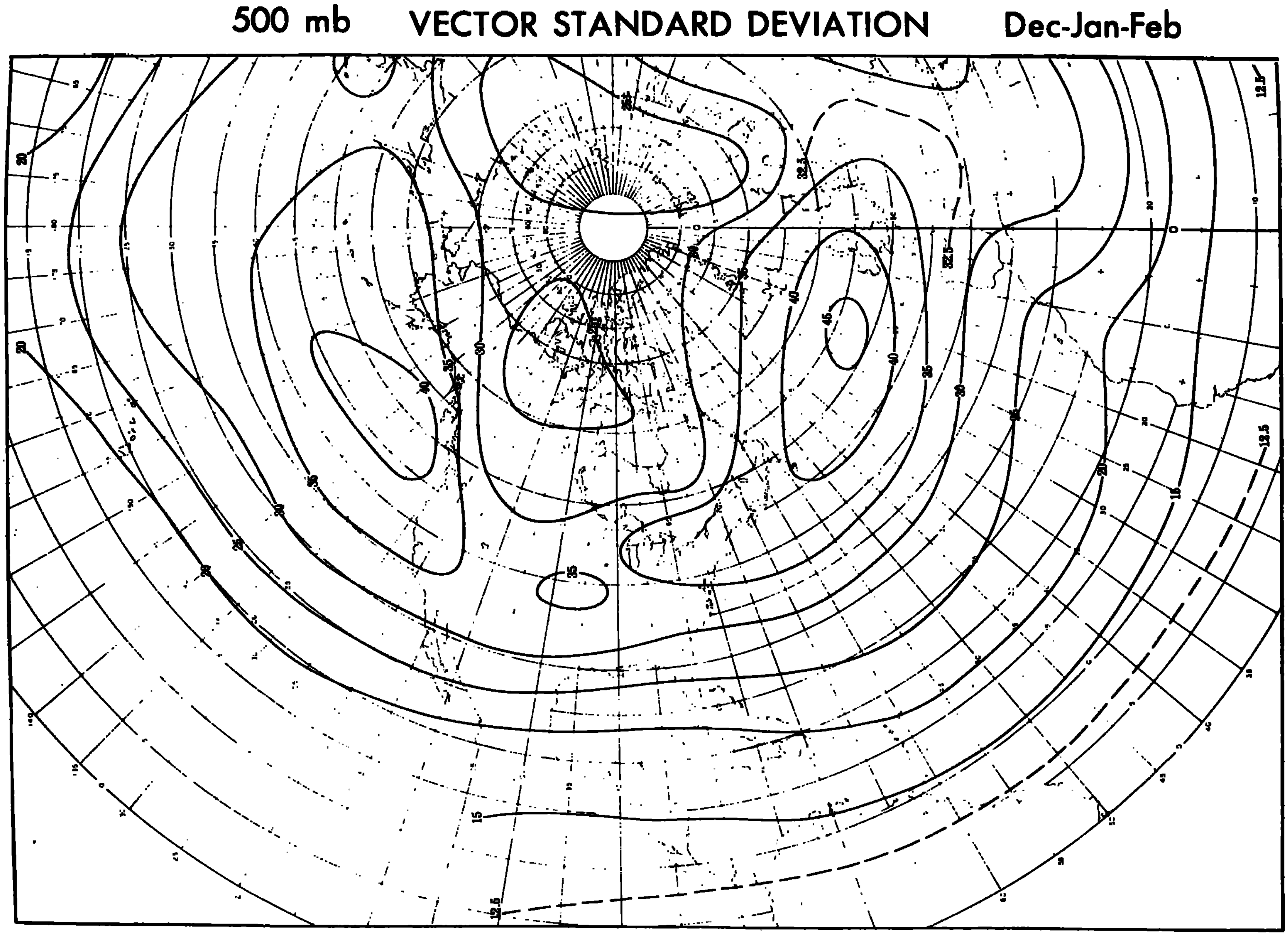

FIGURE 4. Vector standard deviation of the 500-mb wind in winter [knots] (adapted from Crutcher, 1959).

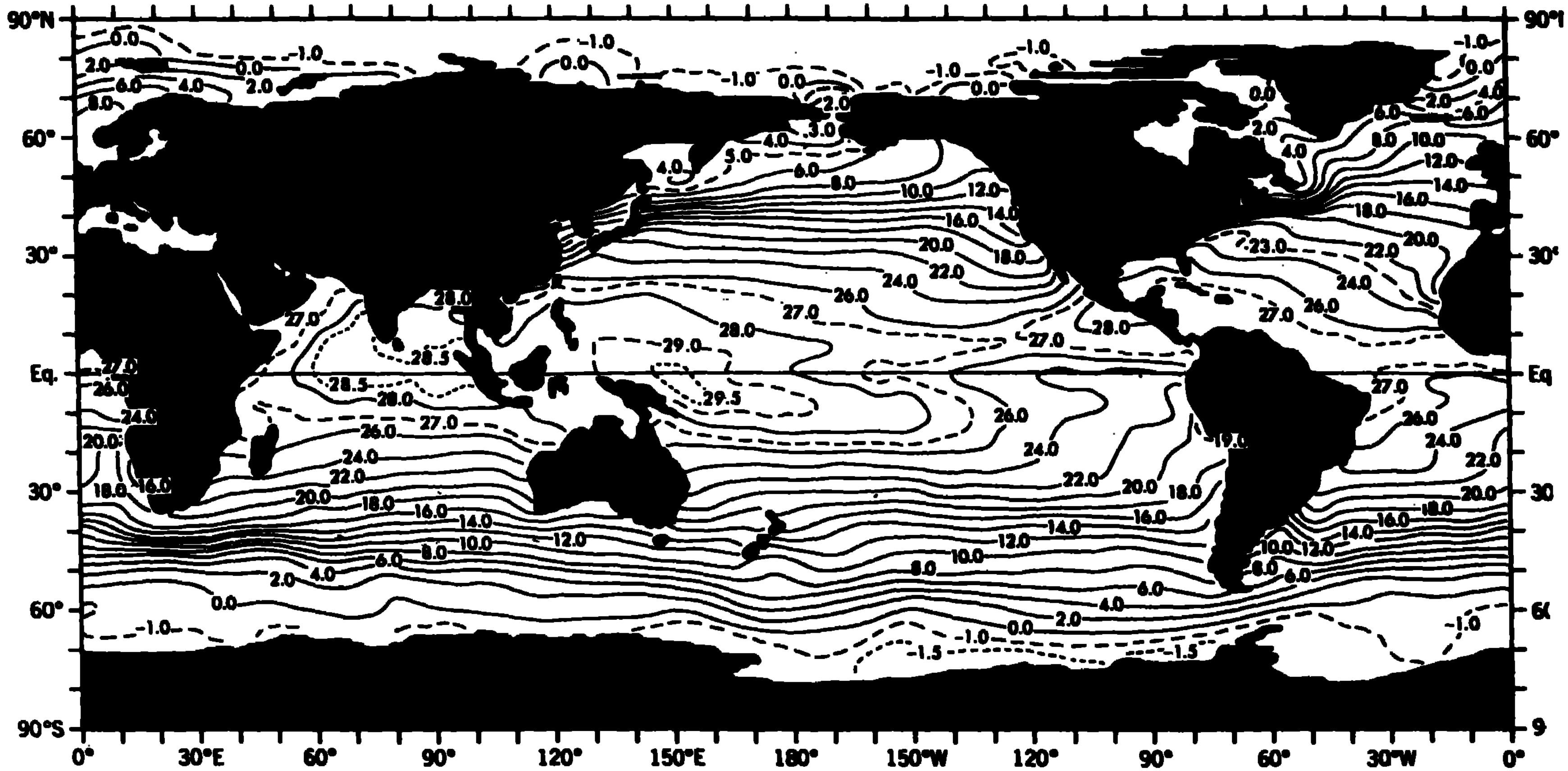

FIGURE 5. Global map of average SST isotherms for the month of January [°C] (from Levitus, 1982.

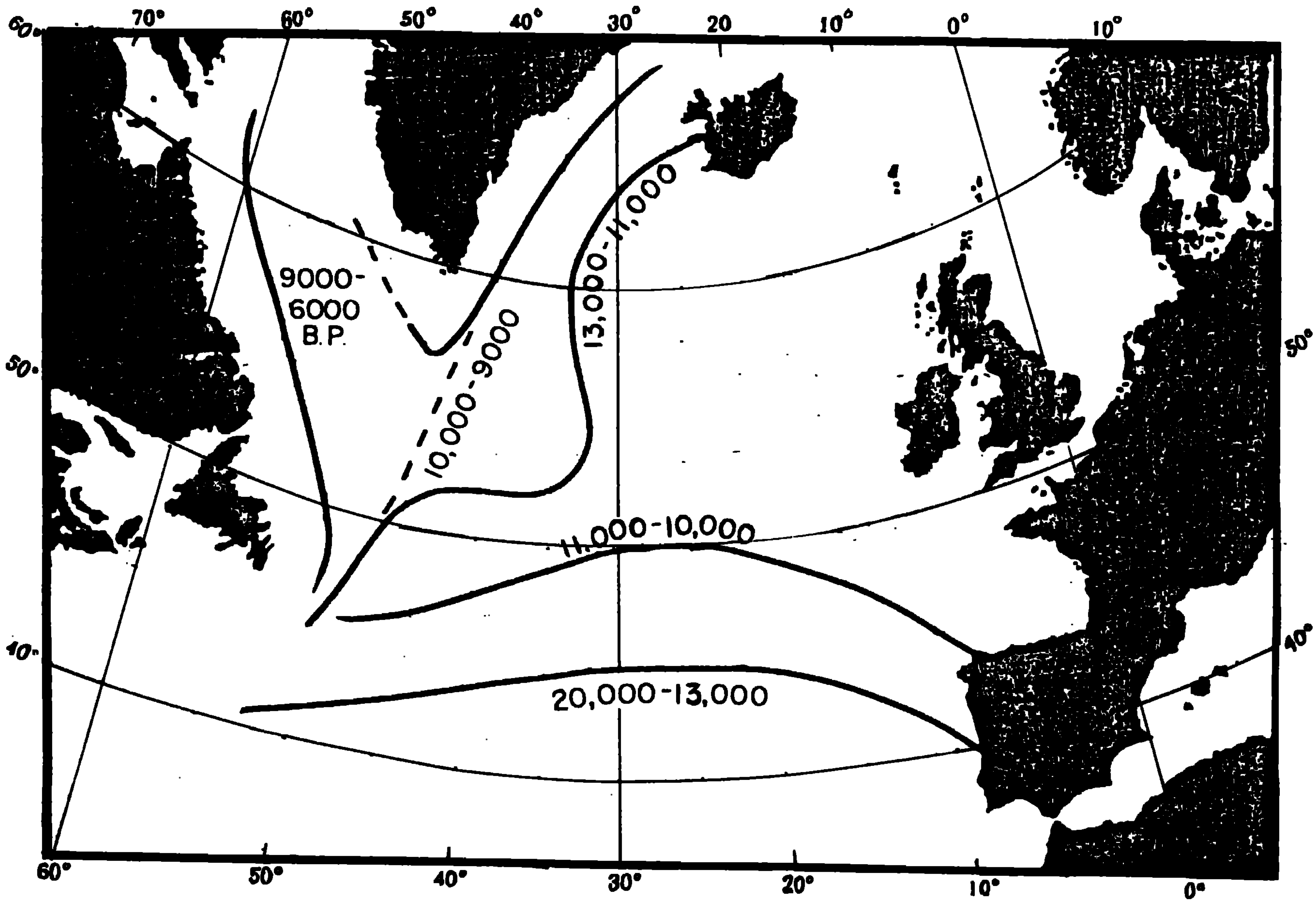

FIGURE 6. Locations of the polar front of the North Atlantic at various times during the most recent deglaciation as determined from ocean sediment cores (adapted from Ruddiman and McIntyre, 1981).